Dynamical Climatology

General circulation modelling of the atmosphere.

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DCTN 77

July 1989

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DYNAMICAL CLIMATOLOGY TECHNICAL NOTE NO. 77

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To appear in Proceedings of NATO Advanced Research Workshop
"Climate-Ocean Interaction",
held in Lady Margaret Hall, University of Oxford, 26th-30th September 1988.

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GENERAL CIRCULATION MODELLING OF THE ATMOSPHERE

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Abstract

The principles involved in constructing and testing an atmospheric general circulation model (AGCM) are discussed. Surface fluxes simulated by versions of the Meteorological Office AGCM are presented and assessed against climatological data. Examples are given of the changes in surface fluxes in numerical experiments with CO_2. Finally, the problems of coupling atmosphere and ocean models are discussed and future advances in atmospheric modelling are considered.

1. Introduction

It is appropriate in a Research Workshop on climate-ocean interaction that the first chapter on modelling should be concerned with the atmosphere. The atmosphere interacts with the ocean through the fluxes of heat, momentum and water at the ocean surface. The ocean affects the atmosphere principally through sea surface temperature, but also through changes in surface roughness and sea ice distribution.

It is not the intention here to give a detailed description of atmospheric models and their capabilities as this has been done recently elsewhere (Schlesinger and Mitchell, 1985; Schlesinger 1988). Instead, the philosophy of constructing and using atmospheric general circulation models will be described and illustrated by specific examples. The remainder of

this chapter is laid out as follows. First, the construction of an atmospheric model is discussed, including the parametrization of those processes not explicitly resolved on the model grid. Then in Section 3 the simulated surface fluxes are assessed, using examples from recent versions of the Meteorological Office atmospheric general circulation model (AGCM). In Section 4 some indications of the likely nature of changes in surface fluxes due to man-induced climate change are presented. In the final section a discussion of the current state of AGCMs and their use in coupled ocean-atmosphere simulations is given.

2. Atmospheric Models

In modelling the atmosphere, one has to take into account both the dynamical relationship between resolved large scale variables such as temperature and wind, and the effect of sub-grid scale diabatic processes including radiation, latent heat release and the exchanges of heat, moisture and momentum with the surface. The radiative effect of trace gases, aerosols and clouds, and the effect of latent heat due to changes of state from water vapour to water and to ice, must all be represented. The specification of the characteristics of the land surface, including its roughness, wetness and reflectivity is also important. In this section we consider first the representation of resolved motions, and then the parametrization of sub-grid scale processes.

2.1 Resolved Motions

In a numerical model, the atmosphere is represented by values of wind, temperature and humidity which are held at various levels at each grid point. The Meteorological Office AGCM (see for example Slingo et al. 1989 henceforth referred to as SWS; Smith, 1989) has 11 levels in the vertical and a horizontal spacing of 2.5° latitude by 3.75° longitude (Fig. 1). For the purposes of this paper, a climate model is an AGLM with an interactive ocean. Current climate models (including those used in the Meteorological Office) use a horizontal grid length which is typically double these values. At any time t the state of the atmosphere is described by about 10^8 variables, and the values at some

future time $t+\Delta t$ can be determined using the prognostic equations described below. Δt is of order 10 minutes.

The Met. Office AGCM currently requires 25 hours of elapsed time on a Cyber 205 for each simulated year, whereas the Met Office climate model requires under 5 hours/simulated year.

The prognostic equations are based on the laws of classical physics, and the local rates of change $\Delta Y_i/\Delta t$ at each location $\underset{\sim}{X}$ are given by

$$\Delta Y_i/\Delta t = F_i(\underset{\sim}{Y}, \underset{\sim}{X}, t) \qquad Y_i = u, v, T, q \qquad (1)$$

where u and v are the east-west and north-south components of wind, T is temperature, q is humidity, F_i is a function of all the variables $\underset{\sim}{Y}$, and their spatial gradients at that location. Equation (1) may be expanded.

$$\Delta Y_i/\Delta t = \underset{\sim}{V}.\nabla Y_i + Z + D + N \qquad (2)$$

(a) (b) (c) (d)

where the term

(a) is the change in Y_i due to transport by the flow

(b) represents the effects of pressure gradients and the earth's rotation in the wind equation, and of adiabatic vertical motion in the temperature equation.

(c) are smoothing (diffusive) terms

(d) are the sources and sinks of momentum (Y_i=u, v), heat (Y_i=T) and moisture (Y_i=q).

Two further equations ensure that atmospheric mass is conserved and that the atmosphere is a perfect gas and in hydrostatic balance. The terms on the left-hand side and terms (a) and (b) are often referred to as the large scale dynamics. The equations for u and v may be transformed into two equations for divergence and the vertical component of vorticity, and so may be solved using spectral techniques: otherwise, they are solved using finite difference techniques (see, for example, Schlesinger, 1988). In solving the equations, care must be taken that there are no spurious sinks or sources of heat, moisture or momentum. This is particularly important for models used in climate change experiments, since the sizes of the perturbations typically considered are small (4 Wm^{-2} in the case of doubling carbon dioxide concentrations (Lal and Ramanathan, 1984)). The diffusive terms (c) are required to prevent energy accumulating in the smallest scales resolved by the model grid and producing non-meteorological solutions. These terms can be regarded as representing the effects of small-scale motions not resolved by the grid.

2.2 Subgrid-scale processes

The final term in (2) covers all the small-scale processes not incorporated in (c), including boundary and surface processes, radiation and convection. For some of these phenomena, notably radiation, there exists an accepted theoretical basis, and accuracy is limited by the availability of computer resources and the necessity of using grid-box values of parameters which in reality may vary substantially within a grid box. For other processes, there may be no accepted underlying theory, or the theory may be valid only for limited idealized conditions. The subgrid scale parameterizations may be based on one or more of the following:

a. Simplified forms of the exact equation (if known)

b. Numerical experiments on a much finer mesh than feasible in the GCM.

c. Observations from the atmosphere

d. Laboratory experiments

e. Sensitivity studies made with general circulation models.

To illustrate the diversity of these approaches, the derivation of the radiative and convective parametrizations are considered in more detail.

First we consider the downward flux of longwave radiation at a level z in the atmosphere. The downward flux at a level z is due to emission from higher levels z' which is partially absorbed at intervening levels z". Downward radiation can arrive from any direction μ in the upper hemisphere, at all wavelengths ν in the longwave spectrum, due to emission from several radiatively active constituents i. The equations of radiative transfer are well known (see for example Paltridge and Platt, 1976). Detailed measurements of the absorption of radiation as a function of wavelength have been made for the principal atmospheric absorbers. Thus the accuracy of radiative fluxes under clear skies is limited only by restrictions in computing resources. Since the explicit integration of the full equations for the vertical flux of longwave radiation F(z)

$$F(z) = \int_0^\infty d\nu \int_0^\infty \mu^{-3} d\mu \int_0^z dz' \, \Pi B(\nu, z') \frac{d}{dz} \exp\left(-\mu \sum_i S_i(z'') f_i(\nu, z'' P(z'') dz''\right) \quad (3)$$

(from Rodgers, 1977) includes a double integration over the vertical co-ordinate z, integrations over all angles represented by μ and all frequencies ν and a summation over all spectral lines i, much effort has been devoted to simplifying the equations and devising efficient

mathematical methods of solution. The Met. Office model currently uses an emissivity approximation which involves a single integration over the vertical co-ordinate and a summation over 7 broad bands (Slingo and Wilderspin, 1986).

The radiative properties of water clouds can be derived theoretically provided the droplet-size distribution is known. Ice clouds are composed of non-spherical particles which may vary in shape and orientation, making it difficult to estimate the radiative properties theoretically. The major uncertainties in radiative effects of ice and water cloud arise in determining cloud properties (height, thickness, extent, water content and droplet (particle) characteristics) from grid scale model variables.

Much of the vertical transport of heat, moisture and momentum is the result of vertical motions on scales much smaller than a model grid box such as cumulus convection. Unlike radiative transfer, there is no generally accepted underlying theoretical basis for parametrizing convective motions. However, some general principles can be applied. If the atmosphere becomes statically unstable, the vertical profile will adjust towards the dry adiabat (or moist adiabat if saturation occurs), conserving moist static energy, $C_pT + Lq + gz$. This is the basis of adjustment schemes, such as the moist convective adjustment scheme of Manabe and Strickler (1964) which instantaneously removes convective instability, and the Betts-Miller scheme (Betts and Miller, 1984) which relaxes the vertical profile towards reference profiles defined from observational data. Other schemes (Arakawa and Schubert, 1974; Lyne and Rowntree, 1976, used in current Met Office models) distinguish explicitly between small-scale ascent in convective towers and large-scale subsidence and warming of environmental air. A third approach taken by Kuo (1974) links convection directly to the large-scale convergence of moisture in a vertical column, including that due to surface evaporation, repartitioning the moisture supply between precipitation and moistening the column. In all but the moist convective adjustment scheme, there are explicit tunable parameters which may be chosen by testing the scheme against observed data such

as those from the GARP Atlantic Tropical Experiment (GATE, Lyne et al, 1976) or by carrying out sensitivity experiments using the full model. Shortcomings of using observational data are: (1) the large-scale convergence of heat and moisture is not well known and (2) it is possible to derive a parametrization for one region which is unsuitable for other parts of the globe. A shortcoming of using sensitivity experiments in that errors in other parametrizations may degrade the performance of the convection scheme. Further guidance may be obtained from observations and numerical simulations of individual convective cells.

Note that even when the theoretical basis for a particular process is well understood and appropriate measurements are available, a certain amount of tuning may be necessary. For example, an error in the observed albedo of clouds of 10% or so would be sufficient to produce sizeable errors in the global mean net downward flux of heat at the ocean surface. This in turn would produce substantial 'climate drift' of the AGCM were it to be coupled to an ocean model. In general, 'tuneable' parameters are kept to a minimum, and retain the same value over the globe. Thus, the opportunities to 'tune' models are limited.

2.3 Using an AGCM

Certain features, referred to here as boundary conditions, must be specified before a GCM can be integrated. For an atmosphere-only model, these include carbon dioxide and ozone concentrations, sea surface temperatures and sea-ice extents, orography and time of year. The normal procedure is then to integrate the model from specified initial conditions over the period of interest and look at the long-term statistics of the simulation. In the case of a climate change experiment, a second simulation is made changing only the relevant boundary conditions, (for example carbon dioxide concentration) and the long-term characteristics of the two simulations are compared. Note that even with fixed boundary conditions, the simulation will vary in time due to the inherent variability of climate. Hence it is

necessary to perform statistical tests on the differences between the control and perturbed simulations in order to establish that the probability that such differences could have arisen by chance is small.

The length of simulation will depend on a variety of factors. First, the time taken for the model to adjust to the prescribed boundary conditions, which will vary from a week or so for prescribed sea surface temperature experiments to decades or even centuries in the case of coupled ocean-atmosphere models. Second, the length of time for which prescribed boundary conditions are valid will also vary, from a month or so for sea surface temperature anomalies to thousands of years in the case of orography. Third, the integrations must be sufficiently long (or numerous) to allow the statistics of the simulated climate to be defined. In practice, the length of a simulation is often limited by the computational resources available.

In the above approach, there are certain underlying assumptions whose validity may affect the relevance of the results to the real climate system.

(a) A "mean" or quasi-equilibrium state exists in nature and in the model. The climate of the earth has varied considerably over a wide range of timescales. Thus in the real world at least, we do not know if statistically stationary climatic states exist. The problem of the stability of climatic states has been addressed using "simple" models (see for example Fraedrich 1978, Ghil and Childress, 1987) and will not be considered further here.

(b) The "mean" state is unique, and independent of the initial state. Lorenz (1975) suggested that the earth's climate may have more than one stable state, and that transitions between these states may be rare (that is climate is almost "intransitive"). It follows that the choice of initial conditions may determine which of the stable states evolve. For example, Wetherald and Manabe

(1975) found that a simulation made with an idealized climate model which commenced with an ice covered surface remained ice-covered, whereas when the initial conditions were free of ice, only polar latitudes became ice-covered.

(c) The model converges, and converges to the correct solution. Although mathematical models of climate are based on sound physical principles, this is not sufficient to guarantee that the model will produce the correct solution. We cannot check the convergence of such complex models using mathematical stability analyses, so the only way to check for convergence is to integrate the model and look at the results.

3. Verification: Surface fluxes

In this section, we consider the problem of verifying an AGCM, with particular emphasis on the surface fluxes which drive the ocean circulation. The validity of the model can be assessed on a variety of timescales against observational data. As climate models can be regarded as a development of numerical weather prediction models, one can specify initial conditions from meteorological analyses, perform short-range forecasts and compare the evolution of synoptic features in the model with those observed (for example, Carson and Cullen, 1976). A similar approach can be taken using long-range forecasts (Mansfield, 1986), though after a period of a week or so, the correspondence between individual simulated and observed features will be lost, and the forecasts of necessity become statistical rather than deterministic.

The traditional method of validating atmospheric GCMs has been to compare the simulated means with climatological data (Manabe and Holloway, 1975; Mitchell 1983). For example, an estimate of the annual mean net flux of heat from the atmosphere into the ocean based on observational data (Figure 2a) may be compared with estimates derived from two different versions of the Meteorological Office AGCM (Figure 2b, c). The climatological data indicate that the ocean is generally heated in low latitudes and cooled in high latitudes as one would expect. In the earlier version of the AGCM

(Figure 2, middle panel) there are large areas of the tropics where the ocean is losing heat, as along the Equator in the Central Pacific, and the Atlantic around 15° of latitude. Because the sea surface temperatures in the AGCM are prescribed, the errors in the surface fluxes did not generate sea surface temperature errors that fed back into the atmospheric circulation, and so did not degrade the simulation significantly. However, the faults evident in Figure 2b would produce large errors if the AGCM were coupled to an ocean model.

The large areas of cooling in Figure 2b were found to be associated with the occurrence of excessive low cloud which reduced the solar radiation reaching the surface. A much improved simulation of the surface heat balance (Figure 2c) was obtained when the treatment of shallow convection was modified (SWS). There are still discrepancies between the observational estimates and the simulated data. However, there are sizeable uncertainties in the observed estimates of net surface heating (see Cattle, this volume) so it is not obvious if and where the model is in error. In this example, the reduction in cloud cover in the tropics increased the solar radiation absorbed both at the surface and at the top of the atmosphere. There is a consequent reduction in reflected solar radiation and in planetary albedo. The revised version of the model produces a good simulation of planetary albedo over most of the tropical oceans, as can be verified by comparing against satellite observations (Figure 3). Thus, the modifications lead to an improvement in the simulation of a directly observable quantity. The albedo over the northern continents is excessive, and this problem has been addressed by including an explicit cloud water variable in the model (Smith, 1988).

The flux of momentum across the ocean surface is a function of the surface flow and low level static stability. For some time, the westerly flow in northern mid-latitudes simulated by atmospheric models became excessively strong as the horizontal resolution was increased (Manabe et al. 1979: Dyson, 1985). This was evident in older versions of the Meteorological Office AGCM with lower than observed sea level pressure over the Arctic and strong zonal flow near 55° North (Figure 4, middle panel). This error has been attributed to the omission of drag due to gravity waves which are not

resolved on the model grid and which propagate upwards from mountains and hills under suitable conditions. Attempts have been made to alleviate the westerly bias by including an explicit parametrization of the orographic gravity wave drag (Palmer et al, 1986). When included in the Met Office AGCM (Slingo and Pearson, 1987), this parametrization produced a marked improvement in low level flows in the northern hemisphere in winter, although the surface flow near the British Isles, for example, is still stronger than observed (Figure 4, bottom panel).

The surface stress depends non-linearly on the surface wind speed and on near surface static stability. The simulated stress over the lower latitude Pacific Ocean (Figure 5, bottom panel) is in broad agreement with estimates derived from climatological data (Figure 5, upper panels). Both the climatological and simulated surface stresses are derived from the synoptic scale winds. Given the quadratic dependence of wind stress on wind speed, it is possible that disturbances on smaller scales than the observational or model grids could increase the surface stress without altering the grid box mean winds significantly.

Finally, we consider the flux of water from the ocean to the atmosphere. Estimates of the evaporative flux from the ocean can be made using climatological data, and this approach was used in deriving the net heat flux shown in Figure 2a. However, direct measurements of evaporation over the ocean are extremely scarce. Precipitation can be measured directly from ships, but there are some doubts about the accuracy of climatological atlases of oceanic precipitation based on ships observations, especially in the tropics (Mintz, 1981). The distribution of precipitation from the simulation described by SWS is shown in Figure 6, where it is compared with the climatological data of Jaeger (1976). Note that the largest discrepancy between the climatological and simulated data occur over the tropical oceans where the observational data are most suspect.

4. Climate Change

The increases in the concentrations of carbon dioxide and other trace gases since 1860 have been estimated to increase the net downward flux at the

tropopause by 2 Wm^{-2} due to the 'trapping' of long wave radiation, and this may reach 6 Wm^{-2} by the middle of the next century (see for example, Mitchell 1989). As the radiative effects of increases in carbon dioxide and the other trace gases are similar (Ramanathan et al, 1985) preliminary investigations of the climatic effect of trace gases have for simplicity considered simulations with enhanced concentrations of carbon dioxide alone, usually with concentrations doubled. Since a doubling of carbon dioxide concentrations would increase the net heating at the tropopause by 4 Wm^{-2} the size of the perturbation in such sensitivity experiments is not unrealistic. In this section, we consider the changes in surface fluxes and related features in a recent study which investigates the effects of doubling atmospheric carbon dioxide concentrations (Wilson and Mitchell, 1987). Note that there are large discrepancies between the size and distribution of changes in climate simulated by different models (Schlesinger and Mitchell, 1987) so the examples given here should be taken as illustrative rather than definitive. Note also that the climate models used in CO_2 experiments have generally had a horizontal resolution with about 5^o latitude by 7^o longitude (or the spectral equivalent) and so are deficient in the simulation of the present day climate.

Although the radiative perturbation due to doubling carbon dioxide concentrations is relatively well known, the response of the atmosphere to changes in radiative heating is not. The radiative warming may be amplified or reduced by a variety of atmospheric processes. For example, in the absence of feedbacks, the 1K increase in temperature would increase the emission of long wave radiation to space by 3.7 Wm^{-2}, so that doubling CO_2 concentrations would give an equilibrium warming of 1.1K (Table 1). The increase in temperature due to the radiative warming allows the atmosphere to hold more water vapour, which also 'traps' long wave radiation (1.4 Wm^{-2}/K), further enhancing the warming. Simulations carried out using atmospheric models coupled with simple mixed layer models of the ocean indicate that the change in snow and ice cover, and the amount and distribution of cloud also enhance the surface warming. These feedbacks are summarised in Table 1 and are discussed in more detail by Schlesinger (1985).

Wilson and Mitchell (1987) found a marked reduction in low and middle level cloud along the depression belts of both hemispheres, which in summer led to a substantial increase in the flux of solar radiation reaching the surface (Figure 7). Note that the size of the increase in surface heating exceeds 30 Wm^{-2} in places and is not distributed uniformly with latitude, but instead is concentrated in the summer mid latitudes.

Simulated changes due to enhanced CO_2 are in general unevenly distributed over the globe. The atmospheric warming is concentrated in the upper troposphere in low latitudes, due to the enhanced release of latent heat, and near the surface towards the Poles, where the greater static stability confines the warming to lower levels (Figure 8a). Thus, the meridional temperature gradient at lower levels is reduced in the subtropics, producing anomalous easterly flow, and is enhanced at upper levels, strengthening the westerly flow (Figure 8b). There are also statistically significant falls in surface pressure over the arctic and rises over the North Atlantic in winter (Figure 9). Thus, as might be expected from Figure 8b, there is a strengthening of the westerly flow at high latitudes, and a weakening of the mean flow near 40°N, with associated changes in surface stress.

Finally, there are non-trivial changes in the supply of water from the ocean surface. The increase in SST due to enhancing CO_2 concentrations leads to enhanced evaporation, particularly in low latitudes. This is because the saturation vapour pressure of water vapour increases exponentially with temperature, so that for any given warming, the largest increases in evaporation will occur where the surface temperatures are highest. The warming of the atmosphere increases its water content, and enhances the transport of heat and moisture to high latitudes (and theintertropical convergence zone) producing enhanced precipitation. As a result, the supply of water to the surface increases in high latitudes where changes in evaporation are small, and it decreases in the subtropics where increases in evaporation are dominant. (for example, Figure 10).

In summary, the changes in the surface fluxes of heat, momentum and water due to doubling carbon dioxide concentrations are likely to be substantial,

and distributed unevenly over the globe. It is likely that such changes will produce significant changes in ocean circulation.

5. Concluding Remarks

In this note, I have attempted to give an indication of how AGCMs are developed, tested and used. As far as possible, model parametrizations are based on sound physical principles. Inevitably, there are processes that are not well observed or understood which result in some uncertainty in the choice of parametrizations. In such cases, it is important to carry out sensitivity experiments to determine whether or not the alternative representations have a significant effect on the simulated circulation. These sensitivity tests should be applied to both the simulation of present climate and of climate change.

There have been significant improvements in the performance of AGCMs over the last four years, both with regard to the simulation of atmospheric flow, and of quantities of secondary interest (at least to atmospheric modellers) such as surface fluxes over the ocean. In the latter case, the improvement has been in response to the demand from those interested in the wider aspect of climate who wish to run coupled ocean- atmosphere models in order to study climate variations such as El Nino and the Southern Oscillation, and changes in climate due to man's activities.

It is evident that there are shortcomings in the climatologies of derived quantities such as surface heat fluxes. As we have seen in section 3, these climatologies may still be adequate to determine the major errors in the simulated quantities. However, a marked improvement is needed in the accuracy of such climatological data if they are to be used to guide the development of improved parametrizations. As the processes governing (for example) surface exchanges vary from region to region, it will be necessary to have measurements that are accurate on a regional scale if they are to contribute to the improvement of models.

It is unlikely that in the forseeable future, atmospheric models will improve to the extent that errors will become negligible. When an AGCM is

coupled to an ocean model, these errors may be enhanced by feedbacks between the atmosphere and ocean, and further compounded by shortcomings in the ocean model. Hence, it seems that for some time there will remain substantial errors in the simulation of present day climate by coupled models. It is known that errors in the control climate will lead to errors in the simulation of climate change (Mitchell et al., 1987). Hence, one is left with a dilemma when attempting to simulate climate change - should one counteract the errors in the atmospheric and oceanic models by adding in appropriate corrections to the surface fluxes (for example, see Sausen et al. 1988) or should one use the unadulterated models, knowing that the errors in the control simulation will distort the model's response?

In the long term, the best approach is to minimise further the errors inherent in the atmospheric and ocean models. However, there is an ever-present demand from 'decision-makers' (who also may control the funding of research) to produce answers now. Thus, it may be more expedient in the short term to perform experiments with and without corrective measures to see if in fact they make any substantial difference to the simulated changes. One must remember that the present state of the art is the exploration of mechanisms of climate change and we are some way from making detailed predictions of climatic change. There are large discrepancies between the currently available studies of the effect of increased carbon dioxide on climate (Schlesinger and Mitchell 1987; Wilson and Mitchell 1987) which should be resolved before attempting to run definitive coupled model experiments.

Improvements in AGCMs are likely to come from two sources. First, increases in computing power will allow higher spatial and temporal resolution, and a more detailed representation of those physical processes which are well understood. Secondly, there will be a slow improvement in the understanding and modelling of the less well understood elements of the climate system, such as cloud and convective processes. Increases in computing power are inevitable, but advances in the understanding of fundamental atmospheric processes will require considerable ingenuity and extensive well designed field and laboratory experiments. This, and not

computing power, will in the long term regulate advances in atmospheric modelling, especially on climatic timescales.

6. Acknowledgements

Section 2 is largely drawn from a lecture given by the author at an European Economic Community School of Climatology and Natural Hazards on "Climatic variations and impacts; an introduction", held in Florence, 11-18 September 1988. I am grateful to Peter Rowntree and Chris Gordon for providing some of the material presented in Section 3, and for constructive comments on the text for Howard Cattle.

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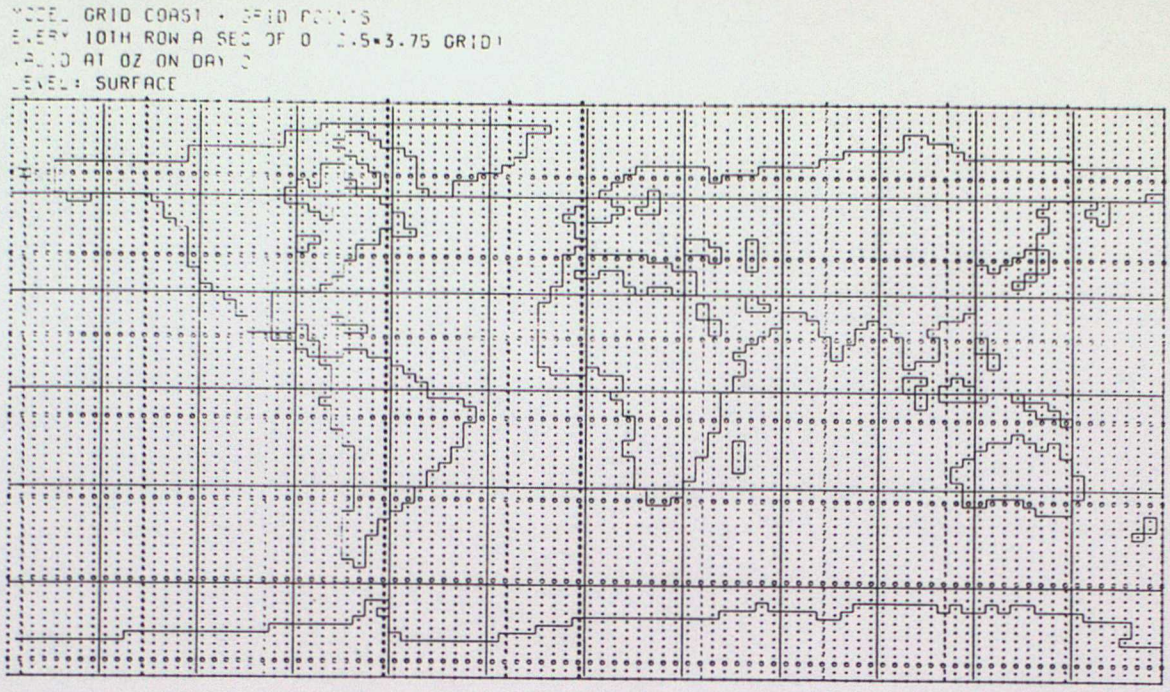
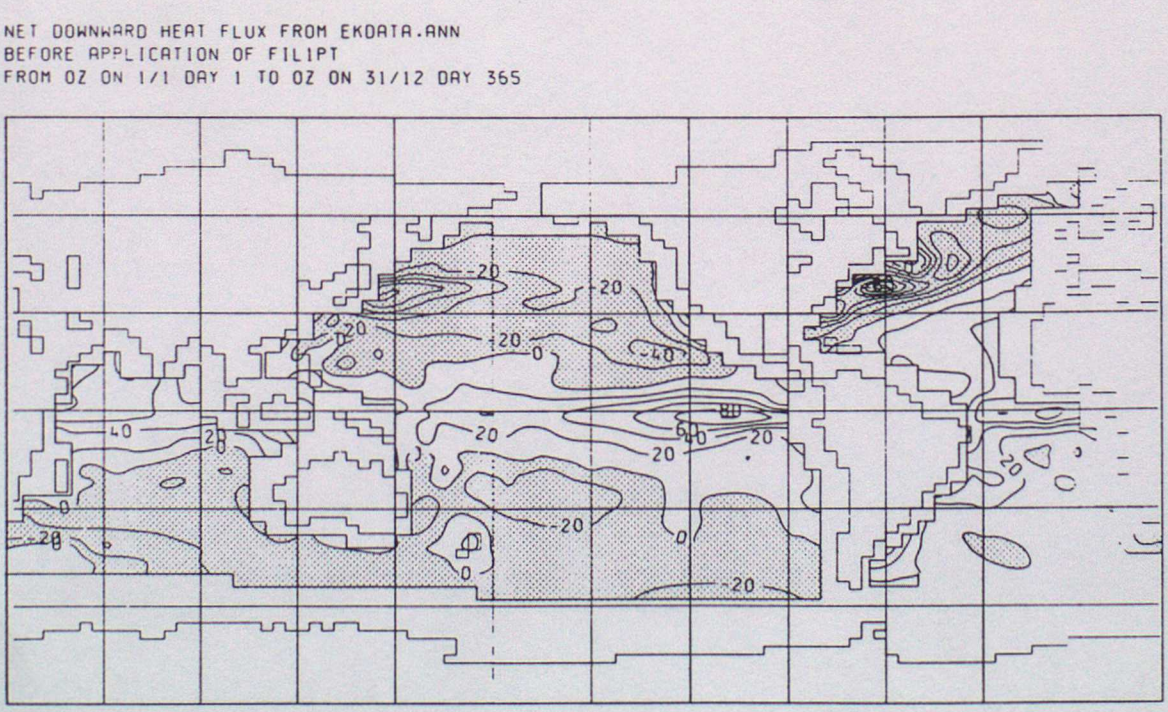
Table 1

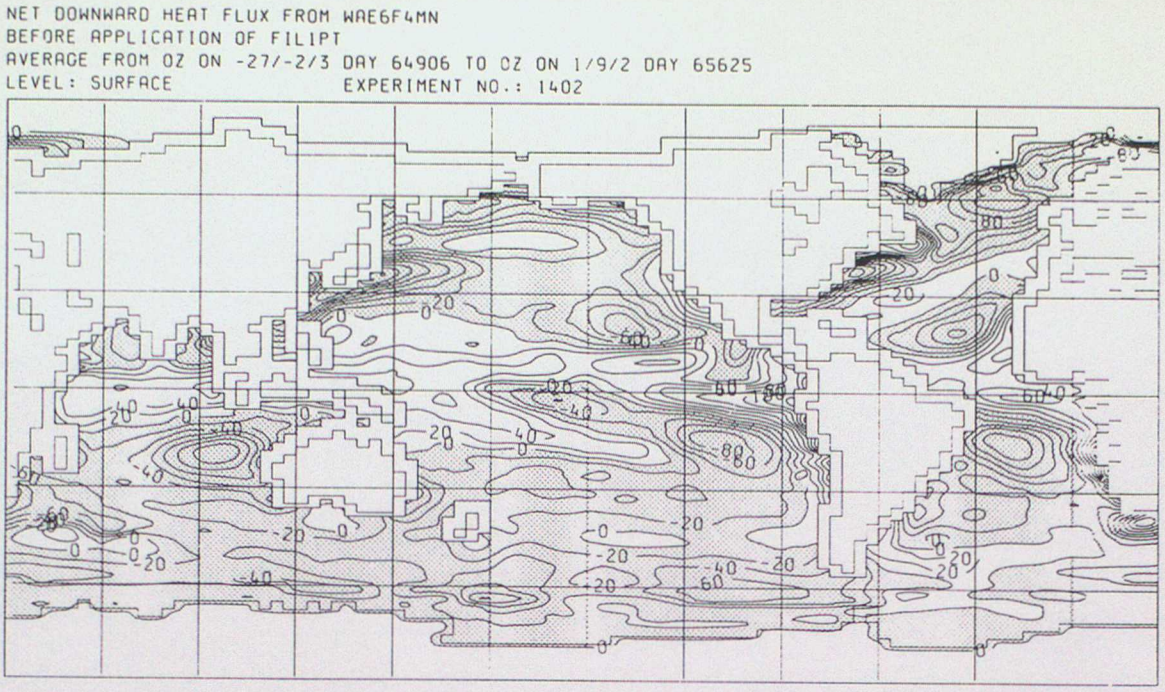
Analysis of Feedbacks and temperature changes due to doubling CO_2, based on typical values from GCM experiments.

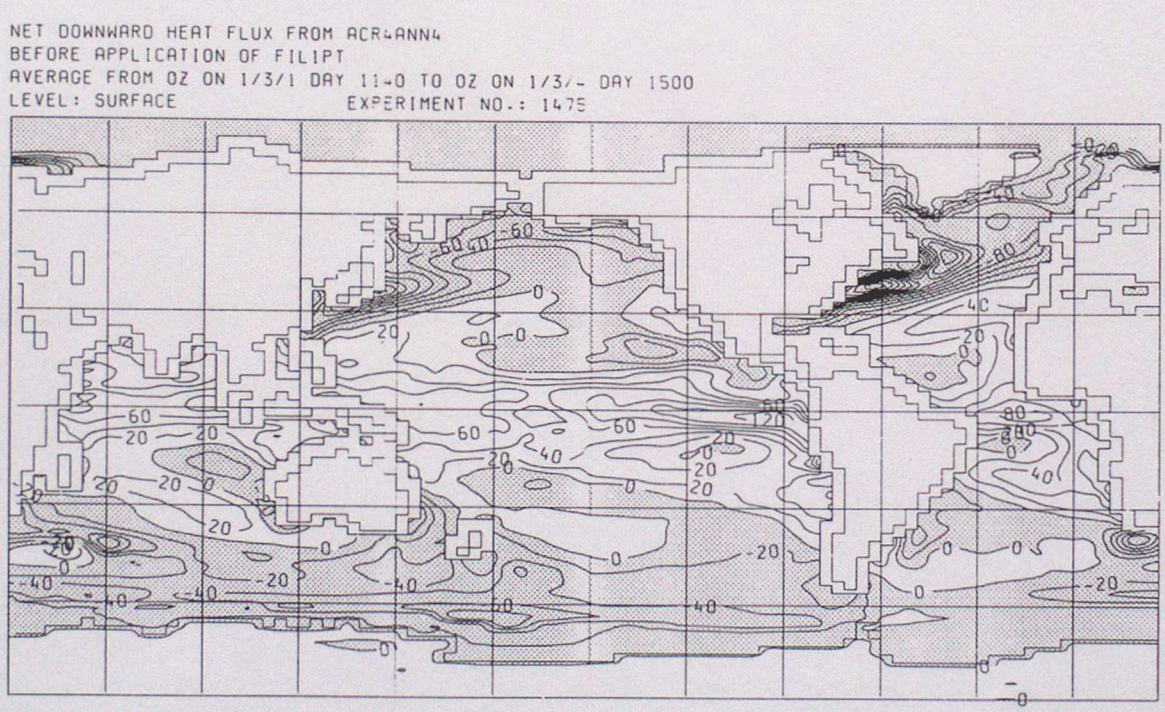
Feedback	Strength ($Wm^{-2}K^{-1}$)	Temperature Change (Cumulative, K)
None	+3.7	1.1
Water Vapour	-1.4	1.7
Ice/Snow	-0.3	2.1
Cloud Amount	-0.9	4.0
Cloud radiative properties	?	?



1. The horizontal grid used in the current Meteorological Office AGCM.


NET DOWNWARD HEAT FLUX FROM EKDATA.ANN
BEFORE APPLICATION OF FILIPT
FROM OZ ON 1/1 DAY 1 TO OZ ON 31/12 DAY 365
-20
-20
-20
-20
-40
-20
-20
-20
-20

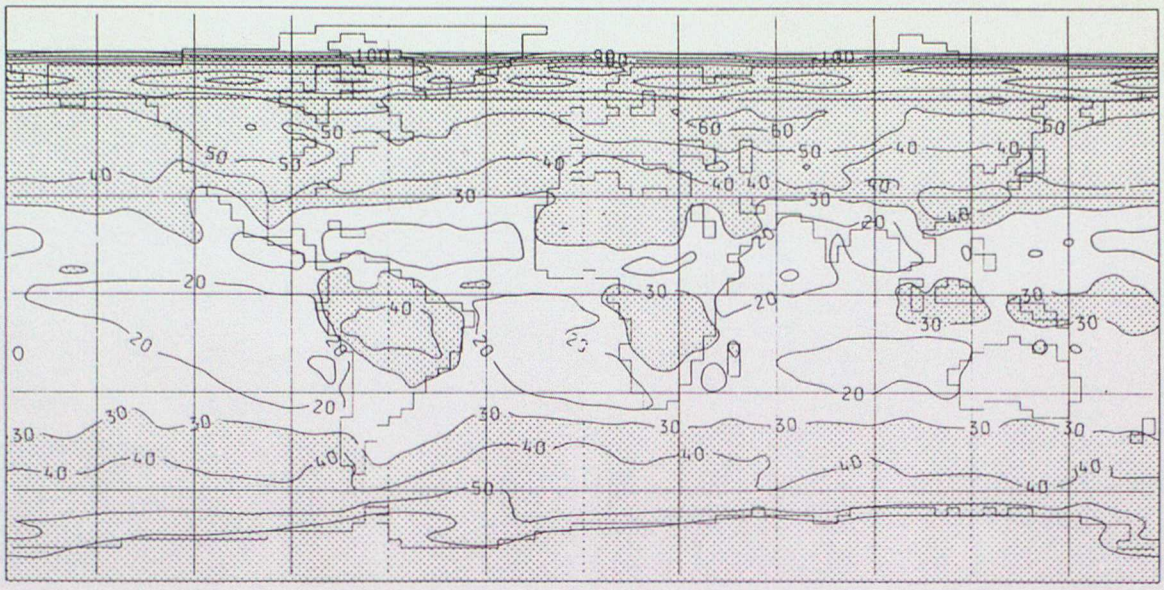




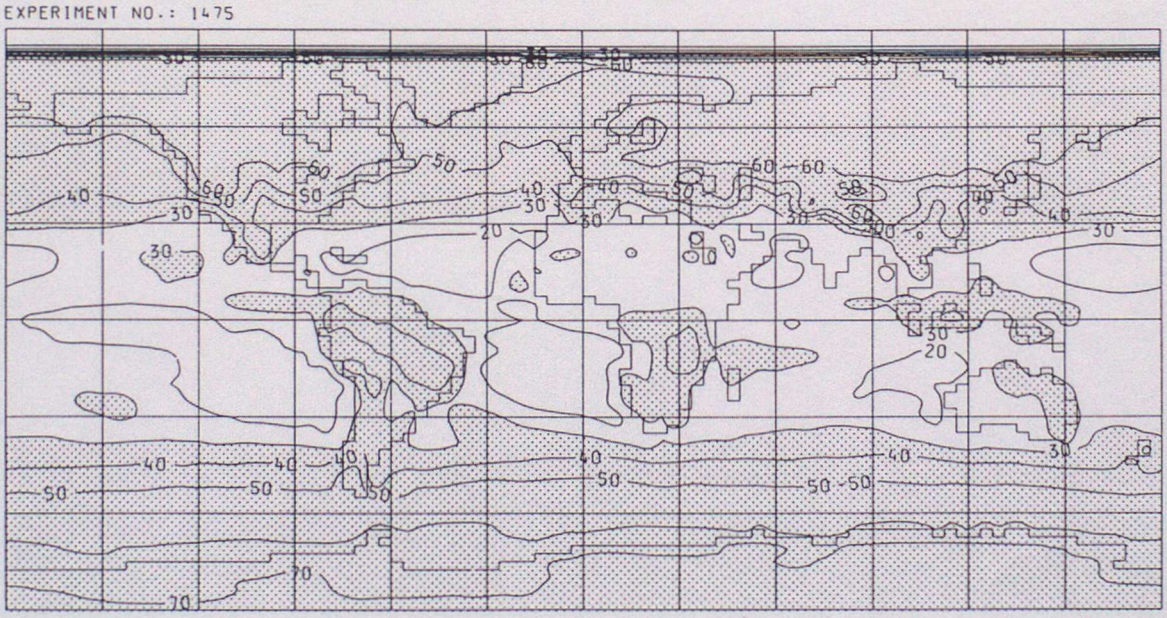
2. Annual mean net surface heating (Wm^{-2}). Contours every 20 Wm^{-2}, and areas of ocean cooling are stippled.

(a) Climatological estimate from Esbensen and Kushnir (1981)
(b) Simulated by an early version of the Met Office AGCM and (c) an improved version (From Slingo et al., 1988).

NFOV ALBEDO FROM NIMBUS-7 DIVIDED BY 1.09
MEAN OVER ALL AVAILABLE WINTERS CONTOURS EVERY 10% SHADED >30
AVERAGE FROM 0.32Z ON 2/12/1978 DAY 336 TO 22.06Z ON 31/12/1978 DAY 365



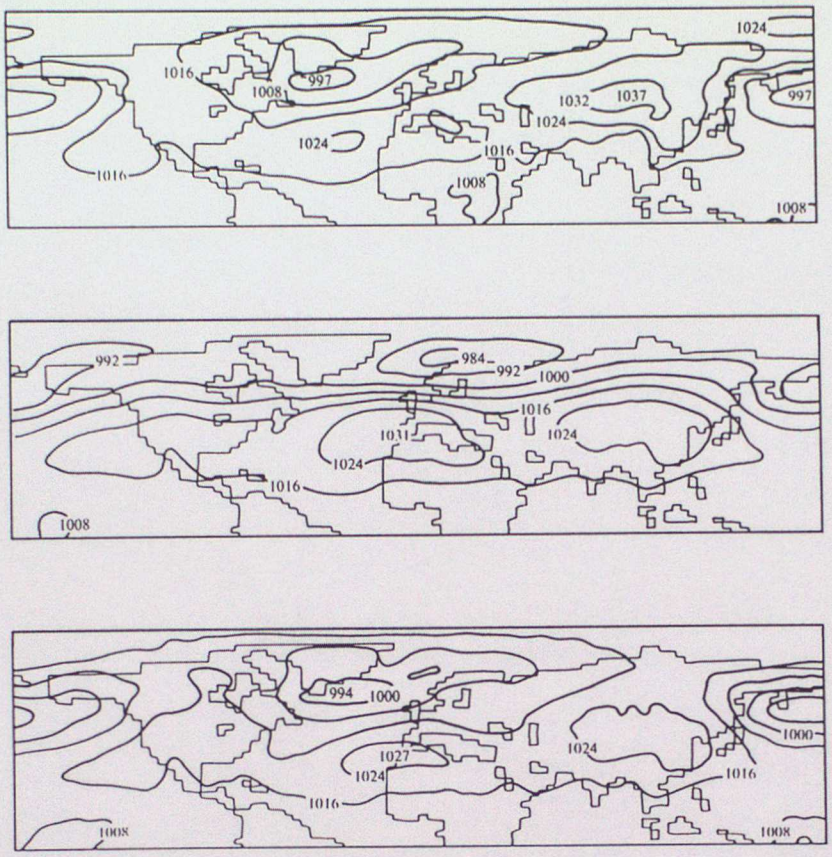
MODEL ALBEDO FOR ACR4WIN4
CONTOURS EVERY 10% SHADED > 30
AVERAGE FROM 0Z ON 1/3/4 DAY 1140 TO 0Z ON 1/3/4 DAY 1500
EXPERIMENT NO.: 1475



3. Shortwave albedo (percent). Contours every 10 percent, and areas greater than 30 per cent are stippled.

(a) Derived from measurements made from NIMBUS7.

(b) Simulated by the Met Office AGM. (From Slingo et al., 1989).

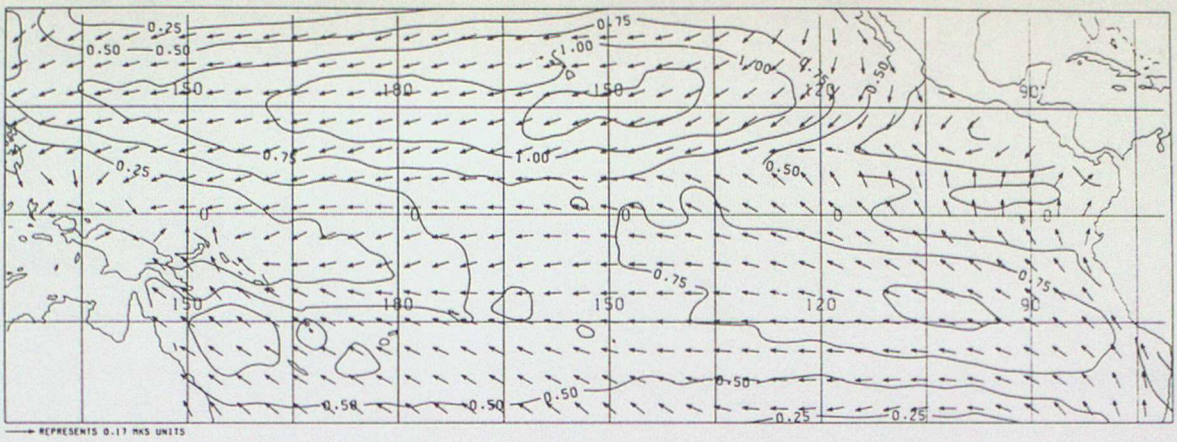


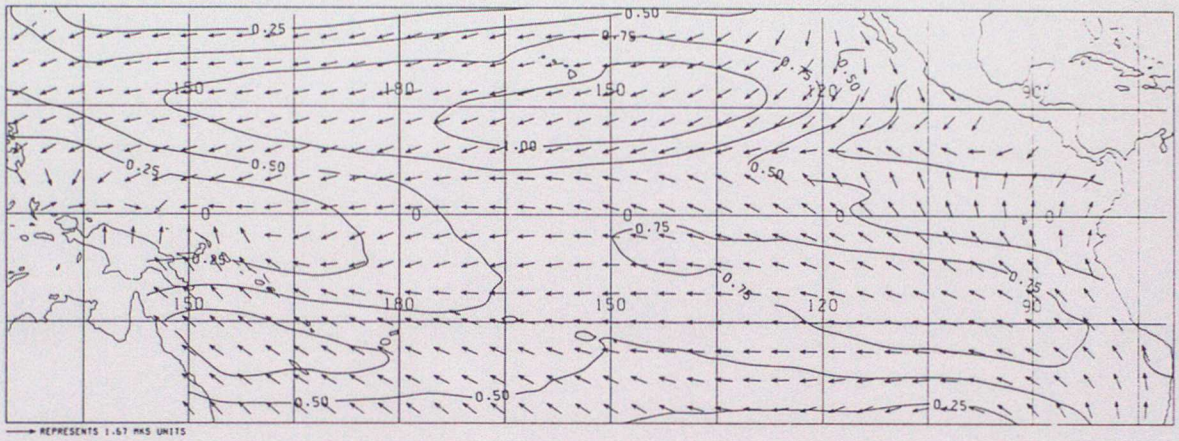
4. Pressure at mean sea level for December, January and February. Contours every 8mbs.

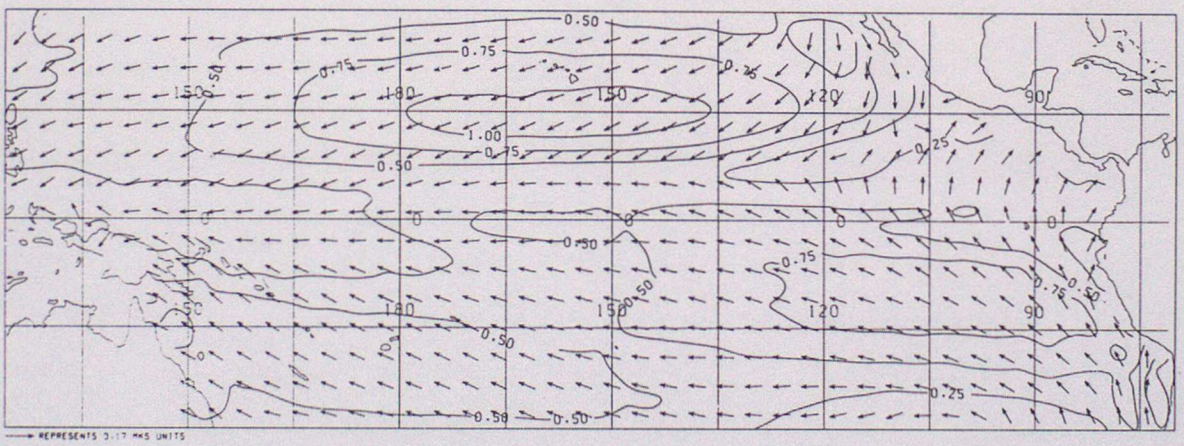
(a) Observed, based on Meteorological Office analyses for 1983-86.

(b) Simulated, from an 8 year simulation without gravity wave drag.

(c) Simulated, averaged over 4 winters with gravity wave drag Slingo and Pearson, 1987, reproduced from Rowntree, 1989)



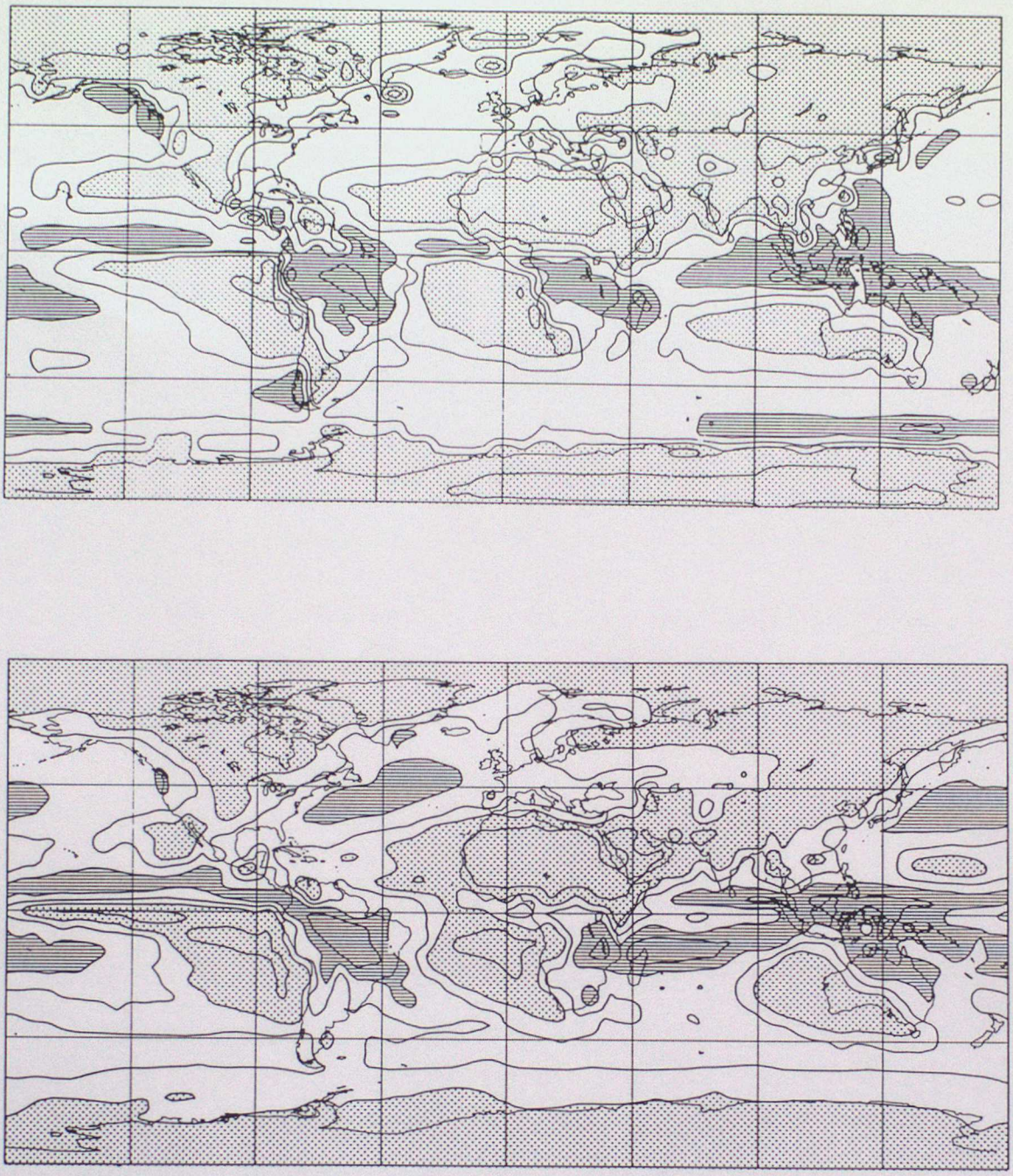




5. Annual mean windstress over the equatorial Pacific. Contours every 0.25 dynes cm^{-2}, with length and direction of the arrows indicating the size and direction of the stress.

(a) Climatological estimate (Hellerman and Rosenstein, 1983).
(b) Climatological estimate (Goldenburg and O'Brien, 1981).
(c) 3 year mean simulated by the Meteorological Office AGCM.

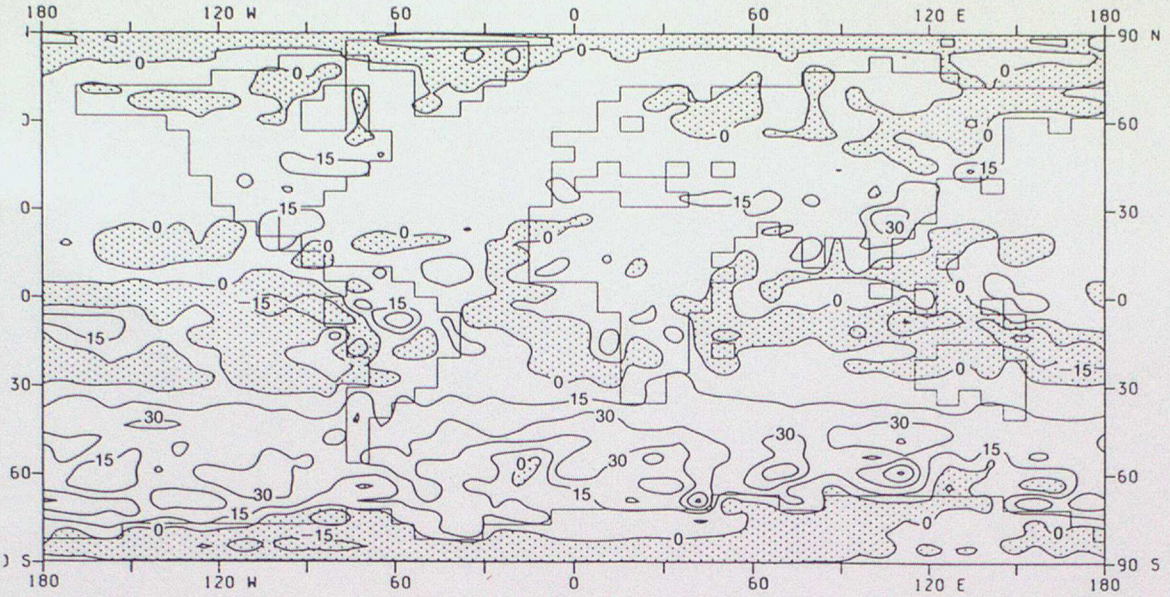
(From Gordon, 1989).



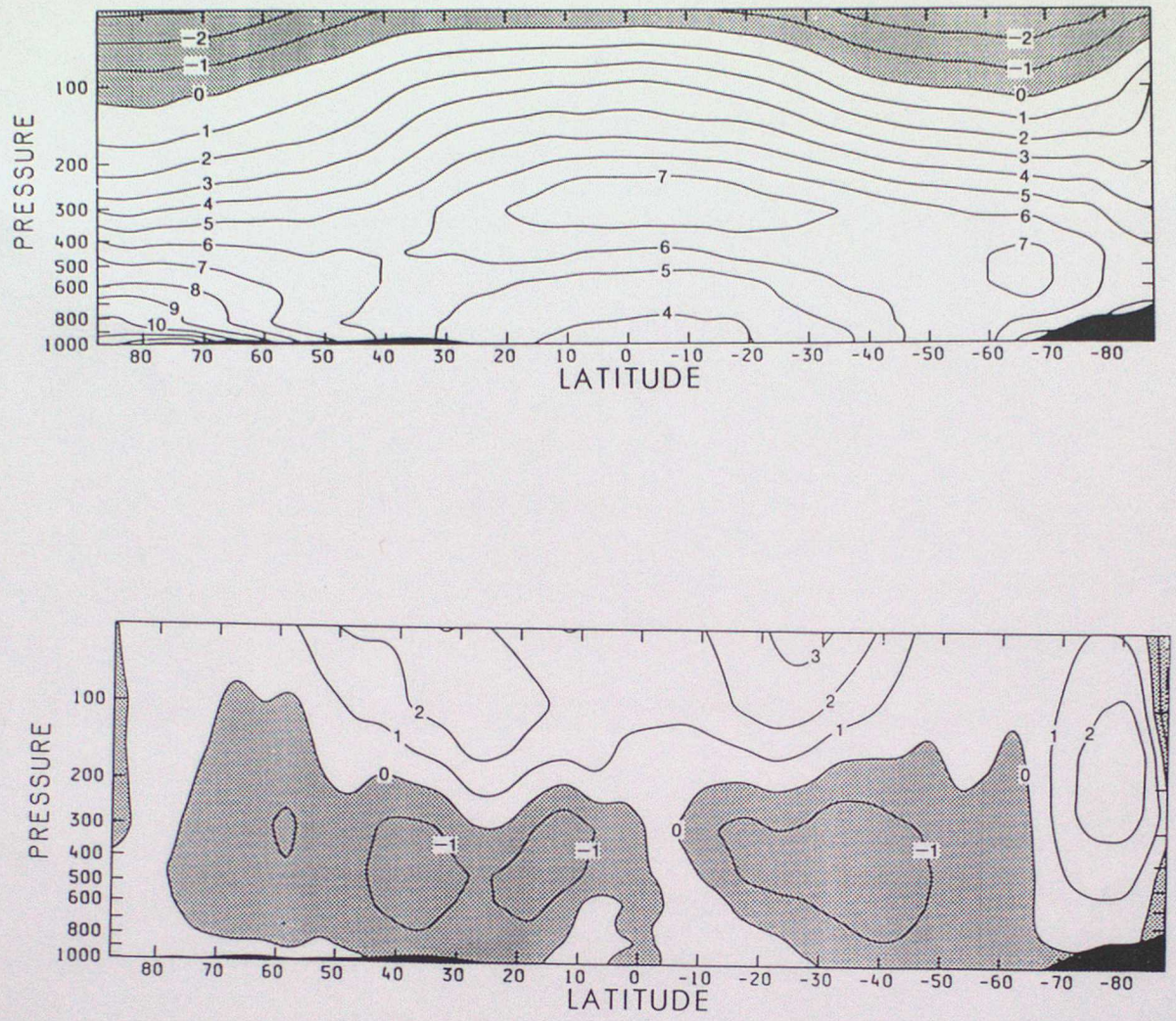
6. Precipitation for December, January and February. Contours at 0.5, 1, 2, 5 and 10 mm/day. Stippled where less than 0.5 mm/day and shaded where greater than 5 mm/day.

(a) From climatological data (Jaeger, 1976).
(b) Simulated.

(From Rowntree, 1989)

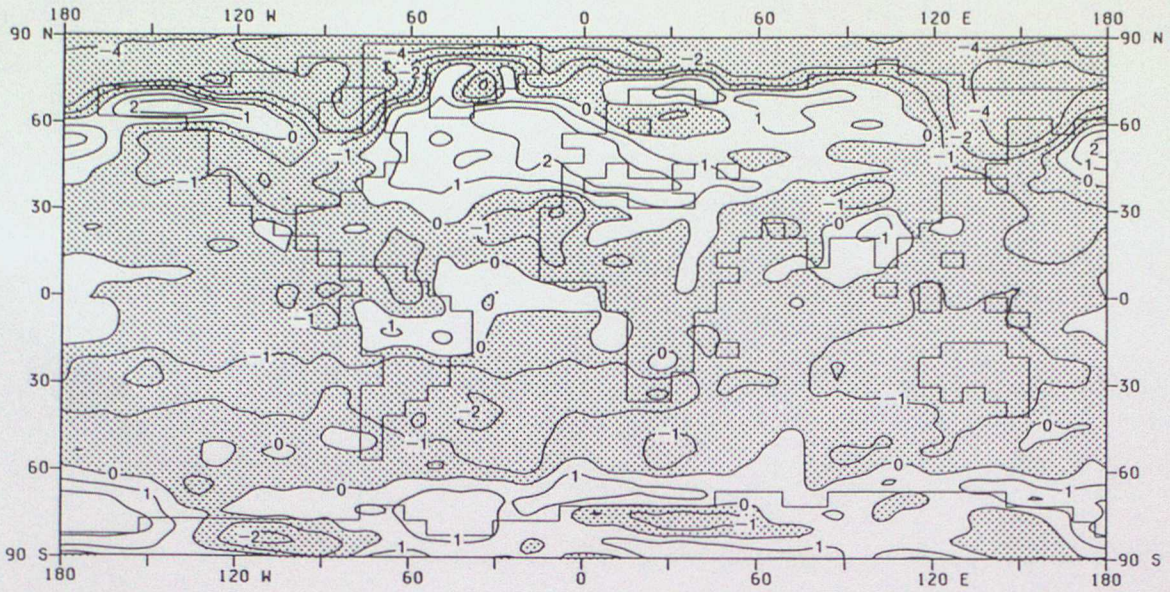


7. Simulated changes in solar radiation reaching the surface due to doubling atmospheric CO_2 concentrations, December to February. Contours every 15 Wm^{-2}, areas of decrease are stippled (from Wilson and Mitchell, 1987).

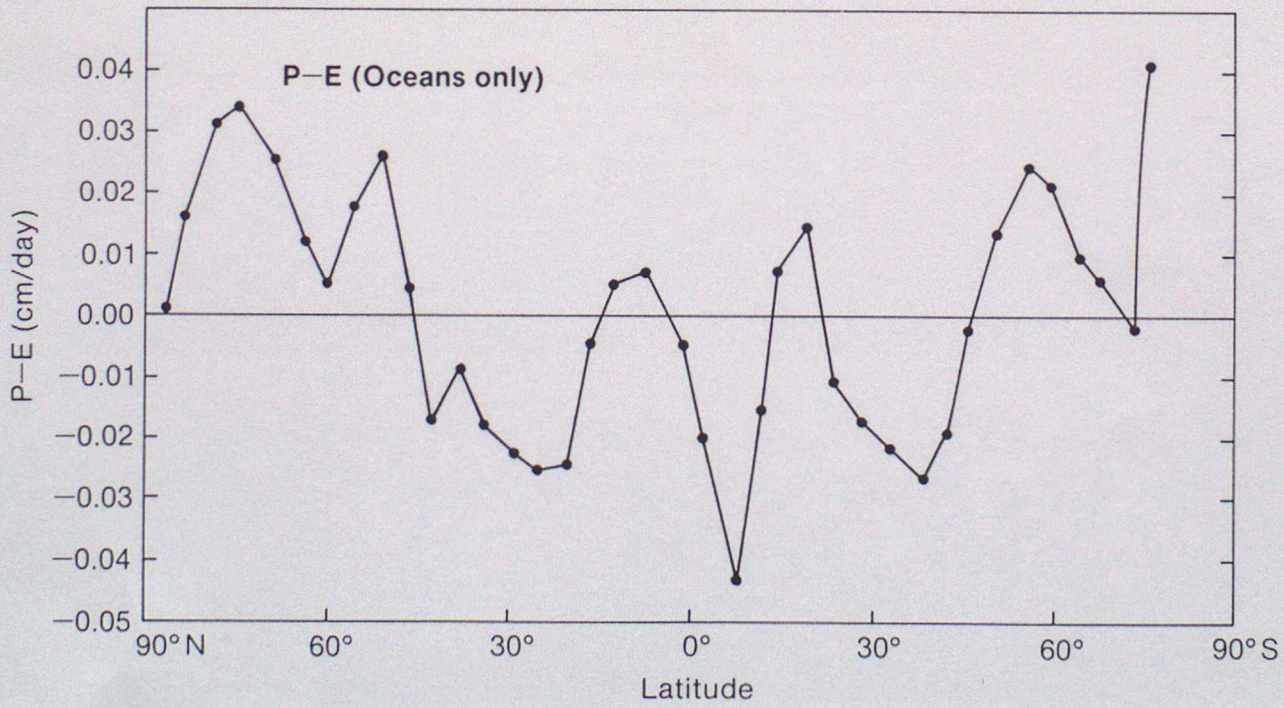


8. (a) Simulated changes in zonally averaged temperature due to doubling atmospheric CO_2 concentrations, December to February. Contours every 1K, areas of decrease stippled.

(b) As (a), except for the west-to-east component of wind. Contours every 1 m/s, regions of weakened westerly flow are stippled. (From Wilson and Mitchell, 1987).



9. Simulated changes in mean sea level pressure due to doubling atmospheric CO_2 concentrations, December to February. Contours at -2, -1, 0 and 1 and 2 mbs, areas of decrease stippled. (From Wilson and Mitchell, 1987).



10. Simulated changes in annual and zonally averaged (precipitation minus evaporation) over the ocean (cm/day) due to quadrupling atmospheric CO_2 concentrations. (Adapted from Manabe and Stouffer, 1980).

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